HADLEY CENTRE
FOR CLIMATE PREDICTION AND RESEARCH

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1. Introduction

The possible role of the land surface in climatic change has been explored in a number of papers (Charney, 1975; Walker and Rowntree, 1977; etc). Aspects of the land surface characteristics which, it has been suggested, may be important include albedo, surface moisture availability and surface roughness. Changes in these can affect evaporation and sensible heat flux, and so modify the temperature and relative humidity and so also the likelihood of rainfall (Rowntree and Bolton, 1983). Each of these surface characteristics can be modified by changes in the vegetative cover. Since many such changes have either occurred recently or in the more distant past, or are thought liable to occur, experiments with three-dimensional models of the general circulation to test all of them would be very demanding of resources, both computational and human. It is therefore useful to consider ways of making at least preliminary assessments in a less time-consuming manner. This paper suggests such a method for a range of land surface changes and gives examples of its use.

In section 2, the relevant equations and parameters are introduced, and some variations in parameters to be expected as vegetation is changed are reviewed. The Penman-Monteith equation is derived in section 3 and applied to estimating the variations of surface heat fluxes with changes in vegetation-dependent parameters. Section 4 considers the effects on evaporation and sensible heat flux of two major changes of vegetation - deforestation and 'desertification'. The impacts of vegetation changes on momentum flux are discussed in section 5. Feedbacks between vegetation and climate are reviewed in section 6, and conclusions are given in section 7.

2. Surface energy balance

2.1 The surface energy balance

The surface energy balance can be written:

$$G = R_N - H - LE \qquad (1).$$

Here, G is the heat flux into the soil; this is small compared with the other terms in (1) in the diurnal mean, though it has a large diurnal variation. In the rest of this section, we will discuss the likely impacts of surface variations on the other terms of equation (1), that is, the surface net radiation R_N, divided into the solar and longwave components

$$R_N = R_{SN} + R_{LN} \qquad (2)$$

and the surface sensible heat flux H and latent heat flux LE.

2.2 Radiative fluxes

2.2.1 Solar radiation

The absorption of solar radiation at the earth's surface can be written as:

$$R_{SN} = (1-\alpha)\ R_S\Downarrow \qquad (3)$$

Here, α is the surface albedo and $R_S\Downarrow$ is the downward solar flux. Both these are spectral quantities, that is they vary with the wavelength of the radiation. This is important because, for example, solar flux is relatively more intense below clouds than with clear skies at wavelengths of less than 0.7µm; at these wavelengths vegetation is particularly strongly absorbing, whereas snow is especially highly reflective.

2.2.2 Range of variation of surface albedo with vegetation

Surface albedo varies from near 0.1 over forests (Oguntoyinbo (1970), Stewart (1971)) to as high as 0.4 over some sandy deserts (e.g. Rockwood and Cox (1978), Kondratyev et al (1982)) and above 0.9 over fresh snow (Kondratyev et al, 1982). With mean clear sky daily solar radiation of 300 W m^{-2} or more in summer in middle and high latitudes, and all year in the tropics, the effects of changes of land surface cover on the energy available for partitioning can approach 100 W m^{-2} - enough energy to warm the atmosphere below a tropopause at 200 mb by over 1 K/day. Even larger changes are possible with snow, as the albedo of forests is little affected by snow once the branches are snowfree, so that albedo decreases of 0.6 with early summer solar radiation can occur with deforestation of high latitudes or of middle latitude mountains.

2.2.3 Longwave radiation

At the temperatures characteristic of the earth's surface, radiation is emitted at wavelengths longer than the approximate upper limit (3μm) of radiation received from the sun. The peak of the emission is at about 11μm, but there is strong emission from about 5 to 50μm. The emissivity of most natural surfaces is near 1 over most of this range (i.e. the surface is almost a black body). The atmosphere emits over a similar range of wavelengths because it has a similar temperature; however, at some wavelengths, the atmosphere is far from being a black body; this is especially so in the so-called water vapour window near 9-12 μm in which there is only weak water vapour absorption and emission, and some strong but narrow trace gas absorption bands, notably due to ozone.

The net absorbed longwave radiation

$$R_{LN} = \varepsilon\Downarrow R_L\Downarrow - \varepsilon\Uparrow \sigma T^4 \qquad (4).$$

Here, there are two emissivities, one appropriate for the spectral composition of the downward radiation, one for the upward, near black-body, radiation from the surface, whose intensity depends on the fourth power of the surface temperature. Outside the water vapour window, $R_L\Downarrow$ originates from levels close to the ground and so usually differs little from the emission from the earth's surface. It is therefore the emissivity in the window which is important for the surface radiation budget.

2.2.4 Dependence of longwave radiation on vegetation changes

Kondratyev et al (1982) reported that the emissivity integrated over the longwave spectrum is about 0.92 for soils and 0.94 for vegetation, while for the water vapour window, the mean for dry soils is 0.96 and for vegetation 0.98. Whilst the relative uncertainties in these figures must be substantial, the suggested value of 0.02 for $\delta\varepsilon$ both for the integrated emissivity (appropriate for upward radiation) and for the water vapour window (more appropriate for the net longwave radiation) is so small as to imply that (with a relatively high net radiation of 100 W m^{-2}), the effect of vegetation on R_{LN} is

$$\delta R_{LN} \approx 2 \text{ W m}^{-2}$$

Emissivities as low as 0.7 have been estimated from satellite data at 9μm for some Saharan surfaces with a large quartz content by Prabhakara and Dalu (1976). Their data suggest that average 9-12μm values probably exceed 0.8 even for these regions, so that changes in R_N due to vegetation changes would be little more than 20 W m^{-2}. Generally, then, the effects of land surface changes on net radiation through emissivity changes can be neglected compared to other changes.

Although these direct effects of a change in vegetation on the net longwave radiation are quite small compared to most other effects we shall discuss, indirect effects due to changes in surface temperature can be substantial; if these are accompanied by changes in the downward longwave flux, they are difficult to allow for in the approach used here; however, where the surface-air temperature difference (δT) increases, by $\Delta(\delta T)$, an adjustment should be made, of

$$-\varepsilon\sigma\Delta(T^4) \approx -4\sigma T^3 \Delta(\delta T)$$

in the net downward radiation R_N.

2.3 Partitioning of the net radiation

2.3.1 Latent heat flux

We consider next the evaporative or latent heat flux. The evaporation for a homogeneous surface can be written in a number of ways:

$$E = \rho\, w'q' = \rho\, C_E\, u\, \delta q = \rho\, \delta q / r_E = \rho\, \delta q \,/\, (r_S + r_A) \qquad (5)$$

Here, w' and q' are small perturbations of vertical velocity and specific humidity respectively, C_E is the surface transfer coefficient for moisture, u and δq are, respectively, the atmospheric wind speed and specific humidity deficit below saturation at a reference height (z_1), commonly 10 m, and r_E is the resistance to moisture transfer. The formal definition of the flux of moisture is given first, then the formulation commonly used in the simpler GCM parametrizations, and then expressions in terms of resistances. Note that for multilayer vegetation, more complicated formulations may be necessary such as that used by Dickinson et al (1986).

As shown in equ (5), the resistance r_E can be separated into the surface resistance r_S and the atmospheric resistance r_A. For vegetation, the former normally represents the stomatal resistance to transfer from moist surfaces within the leaves. It decreases to zero when there is water (dew or intercepted rainfall or snow) covering the stomata and also, in the absence of vegetation, when the soil surface is moist. The resistance r_A represents that between the leaf or soil surface and the atmosphere and is generally taken to be the same as that appropriate for the sensible heat flux discussed below.

2.3.2 Sensible heat flux

Energy is also transported by turbulence in the sensible form:

$$H = \rho\, c_P\, w'T' = \rho\, c_P\, C_H\, u\, \delta T = \rho\, c_P\, \delta T \,/\, r_A \qquad (6)$$

Here, T' is a temperature perturbation and δT is the difference between the surface temperature (T_*) and the potential temperature at z_1.

2.4 Typical values of r_S

As indicated above, the surface resistance, r_S, is zero in the presence of sufficient surface moisture such as when a vegetation canopy is wet. Otherwise, r_S depends on the vegetation type and a number of atmospheric and hydrological variables affecting the supply of and demand for moisture, including soil moisture distribution relative to root development, atmospheric vapour pressure deficit, wind speed, solar radiation and temperature (Jarvis (1976); also see Dickinson (1984) and Stewart (1988) for detailed discussions of the hydrological and atmospheric aspects, respectively).

Even for the relatively simple case of freely transpiring vegetation, observed values of r_S from different sources sometimes conflict; there appears to be agreement on 40 to 60 s/m for growing crops and 80 to 130 s/m for forests in the absence of low temperatures or high vapour pressure deficits (see Perrier (1982), Thompson et al (1981), Buckley and Warrilow (1988)). Grassland values are more variable - from 60 to 200 s/m can be found.

The transpiration cannot exceed the plant's ability to supply water through the roots and stems. This may be seriously limited by soil moisture deficits, which can lead to large increases in r_S. Dickinson et al (1986) modify r_S on the basis of an assumption that the transpiration is restricted to the product of a maximum value (of order 0.5 mm/hour) and a term dependent on soil water potential. This paper mainly considers conditions without limitation of evaporation by soil moisture deficits, as in such conditions estimation of r_S is uncertain.

2.5 Typical values of r_A

2.5.1 Dependence of r_A on roughness length

The atmospheric resistance (r_A) (or $1/C_E u$ from eq. (5)) is a function of boundary layer structure and surface roughness through both the transfer coefficient C_E and the wind speed u. To relate r_A to roughness length z_0 and gradient wind speed V_G, it is necessary to use Rossby similarity theory to find u. The dependence of r_a on z_o and V_G for a range of values of coriolis parameter f and for different values of the Rossby similarity theory parameters A and B is discussed in the Appendix for near-neutral conditions.

Estimates of roughness length (z_o) for different surfaces have been reviewed by Garratt (1977) on the basis of the extensive literature on the subject. Garratt notes that values fall in the range

$$0.02 < z_o/h < 0.2$$

where h is the height of roughness elements. It is common in numerical modelling to assume the roughness lengths for temperature (z_T) and water vapour (z_M) to be equal to z_o; indeed, in some models z_o is assumed to be smaller than z_M. However, Garratt suggests $z_o/z_M \simeq 7$ and some studies propose larger ratios. We shall demonstrate the sensitivity to this assumption by presenting values of r_a calculated first with $z_o = z_M$ and then with $\ln(z_o / z_M) = 2$, close to Garratt's proposal. Table 1, abstracted from the Appendix, shows that r_a is higher with the second assumption, with the fractional change a maximum, nearly a doubling, for the largest z_o (1m) quoted. When we come to apply these results to a possible vegetation change in section 4, we shall see that these are important differences.

Table 1: Variations in r_a (sm^{-1}) for selected z_o and geostrophic wind (V_G) (ms^{-1}) ($f=10^{-4}$ s^{-1})

(a) $z_o = z_M$

z_o (m)	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1
V_G=20	62.6	44.2	28.8	16.4	6.9
V_G=10	120.7	84.8	55.0	31.1	13.0
V_G=5	232.5	162.5	104.8	59.0	24.4
V_G=2.5	447.2	311.2	199.7	111.7	46.0

(b) $\ln(z_o/z_M) = 2$

z_o (m)	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1
V_G=20	73.5	53.7	37.1	23.5	12.8
V_G=10	141.7	103.2	70.9	44.6	24.2
V_G=5	272.9	197.8	135.2	84.6	45.6
V_G=2.5	524.9	378.8	257.5	160.2	86.0

2.5.2 Typical values of z_o

Examples of changes in z_o which are liable to occur are those due to (a) deforestation (or afforestation) involving a conversion from forest to pasture (or vice-versa), and (b) effects of overgrazing and treefelling leading to a change from savannah to sparse grassland or semi-desert. Typical values of z_o for these surfaces are discussed below.

(i) Forest. Tropical forest z_o values of 0.9 to 5m are estimated by Garratt (see appendix to his paper). Note that the largest z_o values in his table (4-5m) are for 30-70% coverage.

(ii) Pasture. Because its height varies widely (from 1cm to 65cm in Garratt's Table 2 with values exceeding 1m for some tropical types (Garratt (1977, p. 19), grassland has a large range of z_o. In our present context of changes over wide areas, we are not concerned with the lowest values of z_o which are for mown grass, nor with the biggest which are presumably for ungrazed land. Grazed areas would be expected to have heights of order 10cm for which $z_o \simeq$1cm. Analyses (see Garratt's review) of the effects of sparse arrays of objects suggest the effects of the grazing animals would be negligible for realistic densities of animals!

(iii) Savannah. For savannah with grass of ≃50cm and shrubs or small trees of 2 to 8 metres, Garratt suggests $z_0 \approx 12$ to 48 cm.

(iv) Sparse grassland or semi-desert. The lowest z_0 for land surfaces, excluding snow, in Garratt's review is that for deserts. Values as low as 0.02-0.03cm have been observed. These were for barchan dunes (Lettau, 1969) and sunbaked sandy alluvium (Calder quoted by Deacon (1953)). Garratt, however, suggests a value for typical Australian deserts of 0.1cm, similar to Clarke et al (1971)'s figure for very sparse grass, arguably an upper bound for non-stony deserts. Note that Sud and Smith (1985) used 0.02cm in a GCM sensitivity experiment to investigate the effects of allowing for the smoothness of deserts.

The above discussion suggests that a range of z_0 from 10^{-4} to 1 m needs to be considered. Table 1 shows that this corresponds to a factor of 10 variation in r_A; there is a strong, roughly linear, inverse dependence on gradient wind speed. As noted earlier, these calculations are for near-neutral conditions. This is generally appropriate as the application is mainly to times of upward energy flux for which departures from neutrality are small. It is relevant that Van Zyl and De Jager (1987) found the Penman-Monteith equation, in which we shall be using these values of r_A, to give good accuracy when compared with lysimeter data, and that the accuracy was not improved by allowing for stability variations.

In section 4, estimates are made of the effects of changes in vegetation due to deforestation and overgrazing. The data above imply that there is a change in z_0 from about 1m for forest to 1cm for pasture. Here the 1m may be conservative, in view of the larger z_0 noted above. For overgrazing, we shall use a reduction from 10cm to 0.1cm. Here both values are arguably too low. However the objective here is to identify broad sensitivities, answering questions such as: - to what vegetation changes are evaporation and sensible heat flux likely to be most sensitive? By taking two changes by the same factor in z_0, but with different starting points, we can isolate the role of the initial roughness in such changes.

3. Dependence of fluxes on surface characteristics

3.1 Use of the Penman-Monteith equation for estimation of surface fluxes

Because of the difficulty of measuring the surface temperature, it is desirable to eliminate it from (6). In the Penman-Monteith approach (e.g. Monteith, 1973), this is achieved by using the surface energy balance:

$$LE = R_N - G - \rho\ c_P\ \delta T / r_A \qquad (7)$$

By writing:

$$\Delta = \partial q_s(T)/\partial T$$
$$\delta' q = q_s(T(z_1)) - q(z_1)$$

so that

$$\delta q = q_s(T_0) - q(z_1) = (\delta T)\Delta + \delta' q \qquad (8)$$

δT (and so T_*) can be eliminated between (5), (7) and (8) to give the Penman-Monteith equation:

$$LE = \frac{\Delta(R_N - G) + \rho\ c_P\ \delta' q/r_A}{\Delta + (c_P/L)(1 + r_s/r_A)} \qquad (9)$$

For some purposes it is more useful to consider the ratio

$$\frac{LE}{(R_N - G)} = \frac{1 + \rho\ c_P\ \delta' q/((R_N - G)r_A\Delta)}{1 + (c_P/L\Delta)(1 + r_s/r_A)} \qquad (10)$$

Note that the equations on which the Penman-Monteith equation is based are commonly used both in forecasting and in general circulation models, so that (9) and (10) can be used to improve understanding of the behaviour of such models. Thus for both real situations and in the modelling context, they can be used to estimate the effects on evaporation of modifications of the vegetation characteristics in respect of moisture availability (by varying r_s), surface roughness (by varying r_a) and surface albedo or other components of the net radiation and surface heat flux (by varying R_N-G).

3.2 Dependence on available energy and vapour pressure deficit

We have noted in section 2 the large variations in albedo and, consequently, net radiation which can result from a change of vegetation. For a typical pair of values of r_a and r_s, Fig 1 shows the effect of varying $(R_N - G)$ with fixed $\delta'q$; for $(R_N - G) < 130$ W m^{-2}, $\mu \equiv LE/(R_N - G) < 0$ so that sensible heat flux would be negative in these conditions. However, with clear skies by day in middle latitude summer or tropical conditions, for which the $\delta'q$ is appropriate, such small $(R_N - G)$, though similar to the daily mean, occur only within 2 to 3 hours of sunrise and sunset. Examination of eq. (10) shows that μ is linearly related to $x \equiv \delta'q / (R_N - G)$. Consequently, the diurnal variation of μ with $(R_N - G)$ can be quite small, since there is a tendency for $\delta'q$, like temperature, to lag only a little behind $(R_N - G)$, with a maximum in the early afternoon (e.g. Shuttleworth et al, 1985). This diurnal asymmetry in $\delta'q$ makes the occurrence of $H < 0$ more likely towards sunset. With typical midday $(R_N - G)$ of 500 W m^{-2}, $LE / (R_N - G)$ is near 0.75, close to what is observed in tropical forest for the daily mean (Shuttleworth et al, 1984).

As $\delta'q/(R_N - G) \to 0$, (10) shows that

$$\mu \to 1 / (1 + (c_p / L\Delta)(1+r_s/r_a)) \equiv 1 / (1 + \varepsilon(1+r_s/r_a)) \quad (11)$$

where ε, so defined, is a function only of temperature. In moist conditions, with $r_s=0$ and $\delta'q=0$, μ takes an asymptotic value $(1+\varepsilon)^{-1}$. Fig 2 and Table 2 show that this asymptotic value, which, as will be shown, also occurs in a number of other contexts, is twice as large for high temperatures (35-40°C) as for temperatures near freezing point.

Table 2: Variation with temperature of Δ ($=dq_s/dT$), ε ($=c_p/L\Delta$) and $1/(1+\varepsilon)$ (i.e. $LE/(R_N - G)$ for the value of r_s at which E is independent of r_a) and $\Delta(1+\varepsilon)$, the denominator in equ. (13).

Temperature (°C)	0	10	20	30	40
Δ (10^{-3} K^{-1})	.28	.516	.908	1.53	2.51
$\varepsilon = c_p/L\Delta$	1.439	.783	.445	.264	.161
$1/(1+\varepsilon)$	.410	.561	.692	.791	.861
Δ $(1 + \varepsilon)$	.68	.92	1.31	1.93	2.91

Differentiation of equ (9) with respect to $(R_N - G)$ reveals that the rate of change of LE with $(R_N - G)$ gives the same expression as in (11):

$$d(LE)/d(R_N - G) = 1 / (1 + \varepsilon (1 + r_s / r_a)) \quad (12)$$

The dependence on temperature and r_s/r_a is displayed in Fig 3. This shows that an albedo change produces a much larger evaporation change with high than with low temperatures, even if $(R_N - G)$ is the same. The decrease in the effect of a change in $(R_N - G)$ on evaporation as surface resistance increases is to be expected since evaporation itself is similarly affected.

Equ. (12) shows that LE always increases with increasing $(R_N - G)$, but that the rate of change as a fraction of that in $(R_N - G)$ is always less than unity. Thus both latent and sensible heat fluxes always increase with an increase in available energy $(R_N - G)$.

In (2.2.4), we considered the possibility of representing the effects of changes in surface temperature on the net radiation, and derived an expression in terms of the change in sensible heat flux H. It is desirable to estimate the magnitude of the adjustment this change in longwave radiation generates for a change in $(R_N - G)$ due, say, to an albedo change. Reference to equ (6) shows that at 25°C, the expression derived in (2.2.4) is about $0.005 r_A \Delta H$ (because of the dependence on ρ as well as T^3, the coefficient varies with the 4th power of temperature). Thus if $\Delta H \approx 0.4\Delta((R_N - G))$, the total change in $(R_N - G)$ after making this adjustment, for a typical r_A of 50 s/m is about 10% less than that expected from the albedo change.

3.3 Dependence on r_A

As already discussed, the atmospheric resistance r_A is a function of wind speed and roughness length z_0, so that as wind speed is itself strongly dependent on z_0 as well as on gradient wind, the gradient wind speed V_G must be specified if sensitivities are to be estimated. The variation of LE /$(R_N - G)$ with r_A is shown in Fig 4(a) for selected values of r_S with values of temperature and $\kappa = \delta'q$ / $(R_N - G)$ appropriate for tropical forests. The r_S values are characteristic of forest vegetation in varying stages of foliage wetness - zero when wet, 100 sm^{-1} when almost dry but freely transpiring.

Of particular interest in Fig 4 is the different response to increasing roughness for low and high values of r_S. This arises from the presence of r_A in both the numerator and the denominator of (10). If r_S is small, the r_A in the numerator dominates the dependence on r_A so that evaporation decreases as r_A increases through, for example, deforestation. On the other hand, for large enough r_S, the r_A in the denominator will dominate and evaporation will increase as r_A increases and limits the importance of the large r_S. Note that as $r_A \rightarrow \infty$, LE / $(R_N - G)$ tends to the asymptotic value $(1+\varepsilon)^{-1}$ already shown in Fig 2 and Table 2.

The critical value of r_S above which E increases with increasing r_A (decreasing z_0) is shown in Fig 5. Note that for typical values of κ of .01 to .03 and for freely transpiring vegetation, this critical value is in the range observed for r_S of 40 to 200 s m^{-1}. For this critical r_S, LE /$(R_N - G)$ again equals the asymptotic value.

The behaviour for zero r_S can be examined by differentiating eq. (9):

$$\frac{d(LE)}{d(r_A)} = \frac{-\rho c_P \ \delta'q \ / \ r_A}{\Delta \ (1 + \varepsilon)} \qquad (13)$$

Table 2 shows the temperature dependence of the denominator; there is a factor of two decrease from 30°C to 10°C, showing that the sensitivity to variations of r_A increases considerably with decreasing temperature. Even greater sensitivity is evident for 0°C, but 10°C is more relevant as it is typical of large areas in middle and high latitudes in late spring and summer when $(R_N - G)$ is comparable to tropical values. Fig 4(b) shows that the larger sensitivity to r_A is also evident for r_S = 50 s m^{-1}, which is typical of growing crops; this was to be expected from the temperature dependence of the critical r_S value in Fig 5, provided κ is not strongly dependent on temperature.

3.4 Dependence on r_S

The previous diagrams have already given an impression of the variation of LE /$(R_N - G)$ with r_S. The variation of this dependence with r_A is clearly shown by Fig 6. Note how much steeper is the cutoff in E with increasing r_S for small r_A - i.e. large z_0 or strong winds. These differences are also in the range of observed variations of r_S, so that a change from forest to growing crops could have substantial effects on LE /$(R_N - G)$, particularly its sensitivity to foliage wetness.

3.5 Additional aspects of the results

3.5.1 Is there an upper limit to temperatures of moist surfaces?

It has been suggested (Newell and Dopplick, 1979; Ellsaesser, 1984) that the temperature of a moist surface is limited to a critical value of about 33°C, above which the latent heat flux LE > $(R_N - G)$, so that downward sensible heat flux is needed to overcome the cooling effect of evaporation. Manipulation of eq. (10) shows that, with $r_s = 0$ for a moist surface, this condition may be written: -

$$\delta'q = q_s(T_A) - q_A > r_A (R_N - G) / \rho L. \qquad (14)$$

Clearly, there is no obvious temperature dependence in this condition. However, for given atmospheric moisture content q_A and $(R_N - G)$, an increase in atmospheric temperature and so $q_s(T_A)$, will lead to $\delta'q$ exceeding the critical value. For typical r_A over water of 140 s/m ($z_0=10^{-4}$ m, Table 1), and typical tropical daily mean $(R_N - G)$ of 150 W m^{-2}, the criterion is $\delta'q > 7 * 10^{-3}$. With boundary layer q of at most $20 * 10^{-3}$ (dew point of 25°C), this limits T_A to just below 30°C, close to the maximum open ocean surface temperature. Note that this application of the Penman-Monteith equation to the ocean assumes G, the heat flux into the surface, to be small, which is not necessarily true for the ocean even for long-period means.

Over land, the midday $(R_N - G)$ of about 500 W m^{-2} (e.g. Shuttleworth et al, 1984) is more appropriate for estimating maximum temperature; this implies a maximum of 35°C, a little above the 33°C suggested by Priestley and Taylor (1972) from observed data. Some discrepancy is to be expected because of the limited duration of such high $(R_N - G)$. The limitation of the argument for an upper limit on surface temperature as greenhouse gases increase is now obvious. In the above, we have assumed an upper limit of 25°C for the atmospheric dew point. As climate warms, there is no reason to assume any such limit.

3.5.2 <u>The concept of equilibrium evaporation</u>

Note that the asymptotic value of LE / $(R_N - G)$ obtained above is also McNaughton and Jarvis (1983)'s equilibrium evaporation, which they derive by seeking the condition that evaporation be constant. However, for the conditions they assumed, a closed boundary layer with prescribed, non-zero radiative input, the boundary layer must be warming and drying and so not be in a true equilibrium.

3.5.3 <u>Evaporation from conifers</u>

As noted previously, the critical value of r_s varies linearly with $\delta'q$ / $(R_N - G)$. Since r_s for certain conifers increases linearly with $\delta'q$ as shown by Roberts (1983), there will be a value of $(R_N - G)$ for which r_s is close to its critical value for all $\delta'q$, so giving constant LE / $(R_N - G)$. According to Roberts, the observed relation is

$$r_s = A\,\delta'q$$

where A is approximately 27 s m^{-1}/(g/kg). With typical growing temperatures of 10-20°C, the critical value of r_s is about 50 s m^{-1} for $\delta'q \approx 0.01(R_N - G)$ (Fig 5). Thus the $(R_N - G)$ required to keep r_s equal to its critical value is (100 * 50 / 27) or 180 W m^{-2}; this is a typical average daytime value during the growing season in the latitude of the observations reviewed by Roberts (e.g. see Kasten (1977)'s data for Hamburg), though, of course, there are considerable variations, both diurnal and seasonal. The evaporation is thus relatively independent of r_A and so, for given vegetation, of wind speed. This may be relevant to the results discussed by Roberts which show that transpiration from conifers has a narrow range of variation both from one day to another and between years and locations.

4. Estimated changes in evaporation due to vegetation changes

4.1 Tropical deforestation

(a) Sensitivity to changes in roughness length (z_o)

The data in section 2.5.2 indicate that a change in vegetation from tropical rain forest to grazed pasture is likely to be associated with a decrease in z_o from about 1m to about 0.01m. Reference to Table 1 and the Appendix shows that this leads to a large increase in the atmospheric resistance r_A. Table 3 shows the effect on $LE/(R_N-G)$ for V_G=2.5 and 10 ms^{-1} and selected r_s in tropical conditions (T=30°C, f=$10^{-5}s^{-1}$) The effects vary linearly with x ($\delta'q$ / (R_N - G)), so it is sufficient to give results for x = 0 and 0.02. Both the assumptions for z_M discussed in section 2.6.1 are used for this case to illustrate the sensitivity to z_o / z_M. The temperature of 30°C is typical of daytime conditions with a dry canopy (e.g. Shuttleworth et al, 1985). However, with a wet canopy, temperatures are likely to be lower, 22 to 25°C being quite common and, as discussed in section 3.3, the sensitivity to r_A variations with r_s = 0 will be larger at such temperatures (by about 30% at 24°C).

Considering first the changes in section (b) of Table 3, these are largest for r_s=0 and for stronger winds, with decreases of up to 3 mm/day. For dry conditions (r_s~100 s/m) on the other hand, changes are quite small for typical x (≈0.02), while for a saturated atmosphere (x=0), there are substantial increases, especially with stronger winds. Deforestation should thus lead to a shift in when evaporation occurs, from wet conditions during and immediately after rainfall with forest, to drier surface conditions, without water on the vegetation but with an ample supply of water to the roots, after deforestation. Some cancellation is clearly to be expected between the two cases for forest (for which typical r_s≈110 sm^{-1} for transpiration with dry foliage), especially where rainfall is followed by relatively moist near-surface air. It may however be important that the immediate recycling of water provided by evaporation from wet vegetation is likely to be substantially reduced as a result of deforestation. Note also that, in this context, McNaughton and Jarvis (1983)'s concept of equilibrium evaporation which takes account of the response to the boundary layer to a change in vegetation may be more relevant to the dry than the wet conditions, as in wet conditions the timescales involved are relatively short.

(b) Sensitivity to albedo changes

Deforestation is also likely to lead to an increase in surface albedo of up to 0.1, e.g. from 0.12 to between 0.17 and 0.22, depending on the type and condition of grass and the extent and type and dryness of any exposed soil. Table 3 (c) shows the changes with an increase of 0.1. The additional increases in LE are the product of the change in (R_N - G), about 23 W m^{-2}, and the value of LE / (R_N - G) for the 'grassland' case. This product ranges from 14 to 23 W m^{-2} for the data in Table 3; in terms of evaporation, this is a decrease of about 0.5 to 0.8 mm / day.

(c) Sensitivity to changes in r_s

Tropical deforestation is also likely to be associated with changes in other land surface characteristics. These include

(i) a probable decrease in the canopy surface resistance in well-watered conditions with dry foliage,

(ii) a decrease in canopy capacity for interception of rainfall, increasing r_s,

(iii) a decrease in root depth, also increasing r_s,

(iv) a decrease in infiltration capacity, increasing r_s.

It is not possible to make any general statement as to what the net effect of these opposing changes might be; estimates should be made for each set of circumstances. Changes in r_s are therefore not considered in Table 3.

Table 3 (a) Values of $LE/(R_N-G)$ for z_0=1m ("forest") and 1cm("grassland")
(assuming $z_O = z_M$)
($f=10^{-5}s^{-1}$, T=303.2K, r_S, r_A in s/m)

(i) V_G=10m/s

	r_A	r_S=0		r_S=50		r_S=100	
		x=0	x=.02	x=0	x=.02	x=0	x=.02
z_0=1m	15.6	.791	1.553	.474	.931	.338	.664
z_0=1cm	63.6	.791	.978	.680	.840	.596	.737
Differences		0	−.575	+.206	−.091	+.258	+.073

(ii) V_G=2.5m/s

	r_A	r_S=0		r_S=50		r_S=100	
		x=0	x=.02	x=0	x=.02	x=0	x=.02
z_0=1m	55.9	.791	1.004	.667	.846	.576	.731
z_0=1cm	233.4	.791	.842	.757	.806	.726	.773
Differences		0	−.162	+.090	−.040	+.150	+.042

(b) Associated differences in latent heat flux ($W\ m^{-2}$) for $(R_N - G)$=150 $W\ m^{-2}$

V_G=10m/s	0	−86	+31	−14	+39	+11
V_G=2.5m/s	0	−24	+13.5	−6	+22.5	+6

(c) Differences in latent heat flux ($W\ m^{-2}$) if also $(R_N - G)$ reduced to 127 $W\ m^{-2}$ by albedo increase of 0.1 from 0.12 to 0.22 (assumes net longwave radiation of 50 $W\ m^{-2}$

V_G=10m/s	−18	−109	+15	−33	+25	−6
V_G=2.5m/s	−18	−44	−4	−25	+6	−12

Table 4 (a) Values of $LE/(R_N-G)$ for z_0=1m ("forest") and 1cm("grassland")
(assuming $\ln(z_O / z_M) = 2$)
($f=10^{-5}s^{-1}$, T=303.2K, r_S, r_A in s/m)

(i) V_G=10m/s

	r_A	r_S=0		r_S=50		r_S=100	
		x=0	x=.02	x=0	x=.02	x=0	x=.02
z_0=1m	29.1	.791	1.200	.582	.883	.461	.699
z_0=1cm	82.0	.791	.936	.702	.831	.631	.746
Differences		0	−.264	+.120	−.052	+.170	+.047

(ii) V_G=2.5m/s

	r_A	r_S=0		r_S=50		r_S=100	
		x=0	x=.02	x=0	x=.02	x=0	x=.02
z_0=1m	104.4	.791	.905	.719	.823	.659	.754
z_0=1cm	301.0	.791	.831	.765	.803	.740	.777
Differences		0	−.074	+.046	−.020	+.081	+.023

(b) Associated differences in latent heat flux ($W\ m^{-2}$) for $(R_N - G)$=150 $W\ m^{-2}$

V_G=10m/s	0	−40	+18	−8	+25	+7
V_G=2.5m/s	0	−11	+7	−3	+12	+3

(c) Differences in latent heat flux ($W\ m^{-2}$) if $(R_N - G)$ also reduced to 127 $W\ m^{-2}$ by albedo increase of 0.1 from 0.12 to 0.22 (assumes net longwave radiation of 50 $W\ m^{-2}$

V_G=10m/s	−18	−61	+2	−27	+11	−10
V_G=2.5m/s	−18	−30	−11	−23	−5	−14

(d) Effect of assuming $z_O = z_M$

Table 4 shows the effects of tropical deforestation estimated assuming $\ln(z_O / z_M) = 2$. The effects without an albedo change are considerably reduced; this reduction is by a factor of 2 for $r_s = 0$ with both wind speeds and generally for the light wind case. These large differences are due to the great sensitivity of evaporation to changes in r_A for small r_A, evident in Fig 4. Because there appear to be good reasons for believing the roughness lengths for moisture and heat flux to be less than for momentum, subsequent results are given only for $\ln(z_O / z_M) = 2$. The effects on the results of changing this assumption in the other cases are similar to those obtained above.

4.2 Middle and high latitude deforestation

The calculations above were made for tropical temperatures and coriolis parameter (f). It is interesting to see the effect of using values more appropriate to middle or high latitudes. Table 5 presents results for temperatures of 283K or 10°C, typical of the growing season in parts of the temperate and boreal forest regions with summer temperatures near the lower limit for growth, such as parts of northwest Europe and Canada. The response for $r_s = 0$ is much larger than for the tropical case. This is to be expected from the temperature dependence discussed in section 3.3. The 20% reductions in r_A also contribute, giving a reduction of about 45%, since dE/dr_A varies with r_A as r_A^{-2} (Eq. (13)). The greater evaporation with a wet surface for forest than for grassland is consistent with observations by Clarke and McCulloch (1979) of much larger evaporative losses in wet conditions from forested than unforested catchments in upland Wales.

Table 5 (a) Values of $LE/(R_N-G)$ for z_O=1m ("forest") and 1cm ("grassland")
(assuming $\ln(z_O / z_M) = 2$)
($f=10^{-4}s^{-1}$, T=283.2K, r_s, r_A in s/m)

(i) V_G=10m/s

	r_A	r_s=0 x=0	r_s=0 x=.02	r_s=50 x=0	r_s=50 x=.02	r_s=100 x=0	r_s=100 x=.02
z_O=1m	24.2	.561	1.667	.294	.874	.199	.592
z_O=1cm	70.9	.561	.939	.428	.717	.347	.580
Differences		0	-.728	+.134	-.157	+.148	-.012

(ii) V_G=2.5m/s

	r_A	r_s=0 x=0	r_s=0 x=.02	r_s=50 x=0	r_s=50 x=.02	r_s=100 x=0	r_s=100 x=.02
z_O=1m	86.0	.561	.872	.447	.695	.371	.578
z_O=1cm	257.5	.561	.665	.517	.613	.479	.568
Differences		0	-.207	+.070	-.082	+.108	-.010

(b) Associated differences in latent heat flux (W m^{-2}) for $(R_N - G)$=150 W m^{-2}

V_G=10m/s	0	-109	+20	-24	+22	-2
V_G=2.5m/s	0	-31	+11	-12	+16	-1

(c) Differences in latent heat flux (W m^{-2}) if $(R_N - G)$ also reduced to 127 W m^{-2} by albedo increase of 0.1 from 0.12 to 0.22 (assumes net longwave radiation of 50 W m^{-2})

V_G=10m/s	-13	-131	+10	-40	+14	-15
V_G=2.5m/s	-13	-46	-1	-26	+5	-15

4.3 Desertification

(a) Sensitivity to z_o

A removal of savanna vegetation and its replacement by sparse grass or semi-desert may, as indicated by the discussion in section 2.5.2, be represented by a decrease in z_o from about 10cm to 0.1cm. Table 1 and the Appendix show that this is associated with an increase in r_A by a factor of nearly 3. The changes in evaporation (Table 6) are of similar character to those for deforestation but substantially smaller - by a factor of 2, even more for r_s=0. This is because of the larger values of r_A. The effects are still substantial, however, amounting to a reduction of 0.7 mm / day with r_s = 0 and x = 0.02 for the strong wind case. As with tropical deforestation, the effects will again be underestimated by Table 6 if temperatures are lower, as is particularly likely in wet conditions.

(b) Sensitivity to albedo

An increase of albedo is also to be expected provided the exposed soil is dry. This would give similar decreases of evaporation (over 0.5 mm / day) to those in the tropical deforestation case. However, the proviso concerning soil dryness implies that such an enhancement could be inappropriate for the r_s = 0 case, so results allowing for this are not included in Table 6.

(c) Sensitivity to changes in r_s

There are likely to be changes in r_s for similar reasons to those stated for deforestation in 4.1 (c). The changes may be more consistently in the direction of an increase in r_s for desertification, since all the processes in 4.1(c) are likely to operate in the same direction. Whereas with savannah, relatively deep roots allow a relatively even supply of water for transpiration, with sparsely covered or bare soil, the evaporation falls off rapidly to low values after rain; Flitcroft et al (1986) report that in such conditions evaporation can be adequately estimated as 4/t mm/day, where t is the time in days since the last rainfall of over 4 mm/day. Thus, r_s, after decreasing during rain, must then increase rapidly; if r_A is about 100 s/m, and x=0.02, Fig 4(a) shows that r_s must increase from about 50 to 500 s/m in one day between t=1 and t=2.

In this case, the adjustment to $(R_N - G)$ to allow for the rise in temperature and upward longwave radiation as evaporation decreases is substantial; for the assumptions listed in Table 6(c), $(R_N - G)$ decreases from 150 to 123.3 W m^{-2}. The change in latent heat flux is consequently enhanced from the 53.8 W m^{-2} expected if $(R_N - G)$ were unaffected to 66.1 W m^{-2}, (evaporation changes of about 1.9 and 2.3 mm/day).

Table 6 (a) Values of $LE/(R_N-G)$ for z_o=10cm ("savanna") and 0.1cm ("semidesert")
(assuming ln (z_o / z_M) = 2)
($f=3*10^{-5}s^{-1}$, T=303.2K, r_s, r_A in s/m)
(i) G=10m/s

	r_A	r_s=0		r_s=50		r_s=100	
		K=0	K=.02	K=0	K=.02	K=0	K=.02
z_o=10cm	48.7	.791	1.035	.652	.853	.554	.725
z_o=0.1cm	110.7	.791	.899	.723	.821	.666	.756
Differences		0	-.136	+.071	-.032	+.112	+.031

(ii) G=2.5m/s

	r_A	r_s=0		r_s=50		r_s=100	
		K=0	K=.02	K=0	K=.02	K=0	K=.02
z_o=10cm	176.0	.791	.859	.747	.811	.707	.768
z_o=0.1cm	408.2	.791	.820	.771	.800	.753	.781
Differences		0	-.039	+.024	-.011	+.046	+.013

(b) Associated differences in latent heat flux ($W\ m^{-2}$) for $(R_N - G) = 150 W\ m^{-2}$

G=10m/s	0	-20	+11	-5	+17	+5
G=2.5m/s	0	-6	+4	-2	+7	+2

(c) Values of LE / $(R_N - G)$ for different r_s with z_o=0.1 cm, allowing for change in longwave radiation due to surface temperature change
(r_A=110.7s/m, x=0.02, G=10m/s)

	$(R_N - G)$	r_s=50	r_s=500
LE / $(R_N - G)$		.821	.462
LE ($W\ m^{-2}$)	150	123.1	(69.3)
H ($W\ m^{-2}$)	150	26.9	(80.7)
LE ($W\ m^{-2}$)	123.3		57.0
H($W\ m^{-2}$)	123.3		66.3

5. Effect of surface roughness changes on momentum flux
The use of Rossby similarity theory in estimating r_A as a function of z_o and gradient wind speed also allows calculation of the surface stress or momentum flux.

Table 7: Wind speed at 10 m and surface stress as a function of z_o for gradient wind speed of 10 m/s

(a) $f=10^{-4}\ s^{-1}$

z_o (m)	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1
u_{10} (m/s)	6.86	6.25	5.43	4.26	2.56
τ($N\ m^{-2}$)	.068	.089	.119	.164	.237

(b) $f=10^{-5}\ s^{-1}$

z_o (m)	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1
u_{10} (m/s)	6.10	5.49	4.69	3.62	2.13
τ($N\ m^{-2}$)	.054	.068	.089	.119	.164

For f typical of middle latitudes ($10^{-4}\ s^{-1}$), a change in z_o from a value of 1m typical of forest to 1cm (short grass) halves the stress despite the more than twofold increase in 10m wind speed. The effects of a larger change from 45 cm to 0.02 cm appropriate for smooth sandy deserts have been studied in a GCM experiment by Sud and Smith (1985) (see section 6.3). In low latitudes, both wind speeds and surface stress are substantially less because for given V_G, u_* varies with fz_o (see equ. (A.6) in Appendix).

6. Feedbacks between vegetation and climate
In the previous sections we have considered the effects of vegetation on climate in terms of 3 parameters, albedo, surface resistance and surface roughness. For completeness, we will now consider possible feedbacks between climate and these vegetation-dependent parameters.

6.1 Albedo
Firstly, for albedo, we have the well-known feedback proposed by Charney (1975) with increasing albedo reducing net radiation both at the surface and at the top of the atmosphere. This requires a change of circulation so as to make good the reduction in heating. In the tropics this is normally achieved through increased subsidence, so drying the troposphere and tending to maintain arid conditions and so the higher albedos. There is also a secondary feedback by which the decrease

in evaporation which we have seen to accompany any decrease in $(R_N - G)$ also tends to reduce the rainfall. On a regional scale, the water balance requires this. GCM experiments support the existence of both feedbacks (Mintz, 1984; Rowntree, 1984).

6.2 Surface resistance to evaporation

Where an increase in resistance to evaporation is due to a loss of vegetation, a positive feedback can be expected because the loss of evaporation can lead to a drier atmosphere and reduction of precipitation, and so, as with the albedo feedback, a maintenance of reduced vegetation. Here, the surface input of energy may not be reduced but, because the upper atmosphere is not heated by latent heat release, subsidence is required there to compensate for radiative cooling (Cunnington and Rowntree, 1986). There is an obvious negative feedback in that the reduction of evaporation helps to conserve soil moisture. There are also radiative feedbacks, positive due to the reduced greenhouse effect of water vapour in the drier atmosphere, positive due to the warming of dry surfaces in the absence of evaporative cooling (as discussed in sections 2.2.4 and 4.3), and negative if the drying of the atmosphere leads to a reduction in cloudiness.

6.3 Surface roughness

There are two mechanisms involved here:

(i) through the effect on partitioning between evaporation and heat flux - we have seen that this has opposite signs for large and small surface resistance. Of particular interest may be the potential for more rapid recycling of water with rough wet surfaces, thus returning moisture to the atmosphere more rapidly from forests than semi-desert or short grass surfaces.

(ii) through the impact on frictional moisture convergence. Increased roughness tends to strengthen cross-isobaric flow and frictional convergence, which by mass conservation leads to more ascent and a greater likelihood of precipitation (Sud and Smith, 1985).

7. Conclusions

Vegetation can affect climate through modification of the heat balance, by changing both the input and also the partitioning of the available energy between sensible and latent heat fluxes. Changes in input are achieved mainly through modifications of albedo; emissivity variations are generally of small importance. However, longwave radiation changes arising from surface warming or cooling can be significant; in particular, surface warming due to reduced evaporation can substantially decrease net radiation so further reducing evaporation. Partitioning is affected by variations in the atmospheric resistance due to roughness changes, and by variations in the surface resistance; the latter may be due to changes in stomatal resistance or to the presence of surface water.

An analysis has been made, using the Penman-Monteith equation, of the likely effects of vegetation changes . An interesting feature of the results is the existence of an asymptotic value of the ratio of evaporation to available energy $(LE / (R_N - G))$, occurring (a) with decreasing vapour pressure deficit or increasing radiation if r_S is zero, or (b) as r_S approaches a critical value , or (c) as r_A approaches infinity. The critical r_S is well within the range of observed r_S, so that care is needed in interpreting the variation of evaporation with roughness: opposite signs of variation will be found for vegetation near wilting point and for vegetation with a wet canopy, while near the critical r_S, a misleading impression of small roughness dependence can be obtained.

The analysis allowed examination of the suggestion that Earth surface temperatures are limited by evaporative cooling. This is shown to be a fallacy in the climate change context, being dependent on an assumption of fixed atmospheric water vapour content.

Evaporation was shown to be much more sensitive to changes in surface resistance (r_s), when the atmospheric resistance (r_A) is small - i.e. for strong winds or rough surfaces.

Changes in evaporation were estimated for particular scenarios. For deforestation, the maximum sensitivity was for wet vegetation. Evaporation decreased by over 1 mm/day in some conditions typical of the tropics. Because of a strong temperature dependence, even larger sensitivities were found for temperatures typical of middle latitudes. This is consistent with observations of larger evaporation from forested than from unforested catchments in wet conditions. With typical r_s for a dry canopy, deforestation could lead to increases in evaporation, though these were generally smaller than the decreases with wet vegetation. These results imply that great care is needed in the representation of canopy wetness in models used to assess the effects of deforestation. An increase in albedo such as might be expected to accompany deforestation was sufficient to eliminate the increases obtained with dry vegetation.

A larger sensitivity was found if the common assumption of general circulation models - that the roughness lengths for momentum and moisture transfer are the same - was adopted. This indicates a need to review this assumption.

For desertification, the effects were smaller than for deforestation because of the larger values of r_A. However, for wet conditions, recycling of water vapour would still be substantially faster for savannah than for desert, especially if there was also an albedo increase.

The roughness changes associated with these vegetation changes were found to change surface stress by about a factor of two for a given gradient wind speed.

There are important potential feedbacks associated with vegetation changes. However, the approach used here does not allow quantification of feedback processes as these depend on changes in circulation, ascent, water vapour convergence, etc. A three-dimensional model is needed for proper study of such feedbacks.

Appendix

The relation between surface roughness and atmospheric resistance

The atmospheric resistance r_A to moisture transfer is defined by (5) with $r_S=0$ as

$$r_A = 1/C_E u. \qquad (A.1)$$

We wish to relate r_A to the surface roughness length z_O. We shall consider here only the near-neutral case, for which most observed data are relevant. The magnitude of the surface stress

$$\tau = \rho C_D u^2 = \rho u_*^2 \qquad (A.2)$$

so

$$r_A = 1/C_E u = (u/u_*^2)(C_D/C_E) \qquad (A.3)$$

where C_D is the drag coefficient for the same level as the wind, typically 10m above the vegetation, and u_* is the friction velocity. Assuming a logarithmic profile

$$\frac{u}{u_*} = \frac{1}{k} \ln\left(\frac{z-d}{z_O}\right) \qquad (A.4)$$

then, for z-d=10m, where d is the zero-plane displacement, since $C_E = k^2/(\ln(10/z_O)\ \ln(10/z_M))$:

$$r_A = \frac{1}{ku_*} \ln\left(\frac{10}{z_O}\right) \ln\left(\frac{10}{z_M}\right) / \ln\left(\frac{10}{z_O}\right) = \frac{1}{ku_*} \ln\left(\frac{10}{z_M}\right) \qquad (A.5)$$

Here k is the von Karman constant, usually taken as 0.4. Because u_* is dependent both on z_O and on the free atmospheric or geostrophic wind speed (V_G), we must use Rossby similarity theory to calculate u_*/V_G from:

$$\ln \frac{V_G}{f z_O} = A - \ln \frac{u_*}{V_G} + \left(\frac{k^2 V_G^2}{u_*^2} - B^2\right)^{1/2} \qquad (A.6)$$

(see e.g. Tennekes and Lumley, 1972, p. 170). Here, f is the Coriolis parameter, and A and B are constants.

The values of A and B for this neutral case were taken from Arya (1975) as A=1.01 and B=5.14. Other values for A and B from Deacon (1973) (1.9, 4.7) and P.J.Mason (personal communication) (1.2, 2.3 - derived theoretically) were also tried. u_*/V_G was calculated by interpolation between values of V_G/fz_O calculated from (A.6) for a range of values of u_*/V_G. Values of r_A for selected z_O were then computed for selected V_G and f. Table A1 presents the results for various z_O, assuming $z_M=z_O$. An obvious result is that the larger z_O is, the more important a change by a given factor becomes.

The comparisons of results for different V_G show only a small (10%) deviation from an inverse linear dependence of r_A on V_G. There is a small decrease in r_A as f increases from $10^{-5}s^{-1}$ (4° latitude) to $10^{-4}s^{-1}$ (44° latitude), but this is only 12½% for $z_O=10^{-4}$m, 20% for z_O=1m. The results for a value of f appropriate for 12° latitude show the variation to be linear with log f. Broadly, then, the ratios between values of r_A for a given z_O change will depend little on V_G or f. The calculations with different A and B are summarised in Table A2. Deacon's and Mason's values give similar results, lower than those with Arya's - by 5% at $z_O= 10^{-4}$m, 10 to 15% at z_O=1m. Our choice of Arya's A and B will turn out to be conservative since the other two pairs of values would give smaller r_A and increased ratios between r_A for different z_O.

As discussed in section 2.5.1, it is probable that z_O exceeds z_M and most of the results in this paper have assumed that $\ln(z_O / z_M) = 2$. Inspection of equ. (A.5) shows that this increases r_A with the biggest fractional increases for the largest z_O. Table A3 shows that r_A is almost doubled for z_O = 1m. The ratios of r_A between different V_G are unchanged.

Table A1: Variations in r_a (sm^{-1}) for selected z_0 with coriolis parameter (f) (s^{-1}) and geostrophic wind (V_G) (ms^{-1}) ($z_0=z_M$)

z_0 (m)	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1
Variation with V_G					
$f=10^{-4}$					
$V_G=20$	62.6	44.2	28.8	16.4	6.9
$V_G=10$	120.7	84.8	55.0	31.1	13.0
$V_G=5$	232.5	162.5	104.8	59.0	24.4
$V_G=2.5$	447.2	311.2	199.7	111.7	46.0
$(V_G=2.5)/(V_G=10)$	3.71	3.67	3.63	3.59	3.55
$f=10^{-5}$					
$V_G=10$	135.8	96.6	63.6	36.6	15.6
$V_G=2.5$	506.7	357.8	233.4	133.1	55.9
$(V_G=2.5)/(V_G=10)$	3.73	3.70	3.67	3.63	3.59
Variation with f					
$V_G=10ms^{-1}$					
$f=10^{-4}$	120.7	84.8	55.0	31.1	13.0
$f=3*10^{-5}$	128.5	90.9	59.4	34.0	14.3
$f=10^{-5}$	135.8	96.6	63.6	36.6	15.6
$(f=10^{-5})/(f=10^{-4})$	1.125	1.139	1.157	1.178	1.201

Ratios for variation with z_0(m)

	10^{-4} -> 10^{-3}	-> 10^{-2}	-> 10^{-1}	-> 1
$V_G=10$, $f=10^{-4}$	1.423	1.542	1.769	2.401
$V_G=10$, $f=10^{-5}$	1.406	1.519	1.736	2.355
$V_G=2.5$, $f=10^{-4}$	1.437	1.558	1.788	2.428

Table A2 Comparison of results with different (A,B) values
($f=10^{-4}$ s^{-1}, $V_G=10$ ms^{-1}) ($z_0=z_M$)

z_0 (m)	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1
Arya (A=1.01, B=5.14)	120.7	84.8	55.0	31.1	13.0
Deacon (A=1.9, B=4.7)	114.1	79.5	51.0	28.5	11.7
Mason (A=1.2, B=2.3)	115.1	79.8	50.8	27.9	11.1

Table A3: Variations in r_a (sm^{-1}) for selected z_0 with coriolis parameter (f) (s^{-1}) and geostrophic wind (V_G) (ms^{-1}) ($\ln(z_0/z_M) = 2$

z_0 (m)	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1
Variation with V_G					
$f=10^{-4}$					
$V_G=20$	73.5	53.7	37.1	23.5	12.8
$V_G=10$	141.7	103.2	70.9	44.6	24.2
$V_G=5$	272.9	197.8	135.2	84.6	45.6
$V_G=2.5$	524.9	378.8	257.5	160.2	86.0
$f=10^{-5}$					
$V_G=10$	159.4	117.6	82.0	52.5	29.1
$V_G=2.5$	594.7	435.5	301.0	190.9	104.4

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Figure captions

Figure 1: $LE / (R_N - G)$ as a function of $(R_N - G)$ for typical tropical conditions (near surface air temperature of 303.2K, r_A of 50 s/m, r_s of 50 s/m and $\delta'q$ of 5 g/kg.

Figure 2: The asymptotic value $(1+\varepsilon)^{-1}$ of $LE / (R_N - G)$ as a function of temperature.

Figure 3: The change in latent heat flux, as a fraction of the change in $(R_N - G)$, as a function of temperature and r_S/r_A.

Figure 4: $LE / (R_N - G)$ as a function of r_A for various values of $x=\delta'q/(R_N - G)$ and r_s with near-surface air temperatures of (a) 303.2K and (b) 283.2K.

Figure 5: The critical r_S below which evaporation increases with roughness length for two values of $x=\delta'q/(R_N - G)$.

Figure 6: Dependence of $LE / (R_N - G)$ on r_S for r_A of 10 and 100 s/m, with near-surface temperature of 303.2K and $\delta'q/(R_N - G) = 0.02$ (g/kg)/(W m^{-2}).

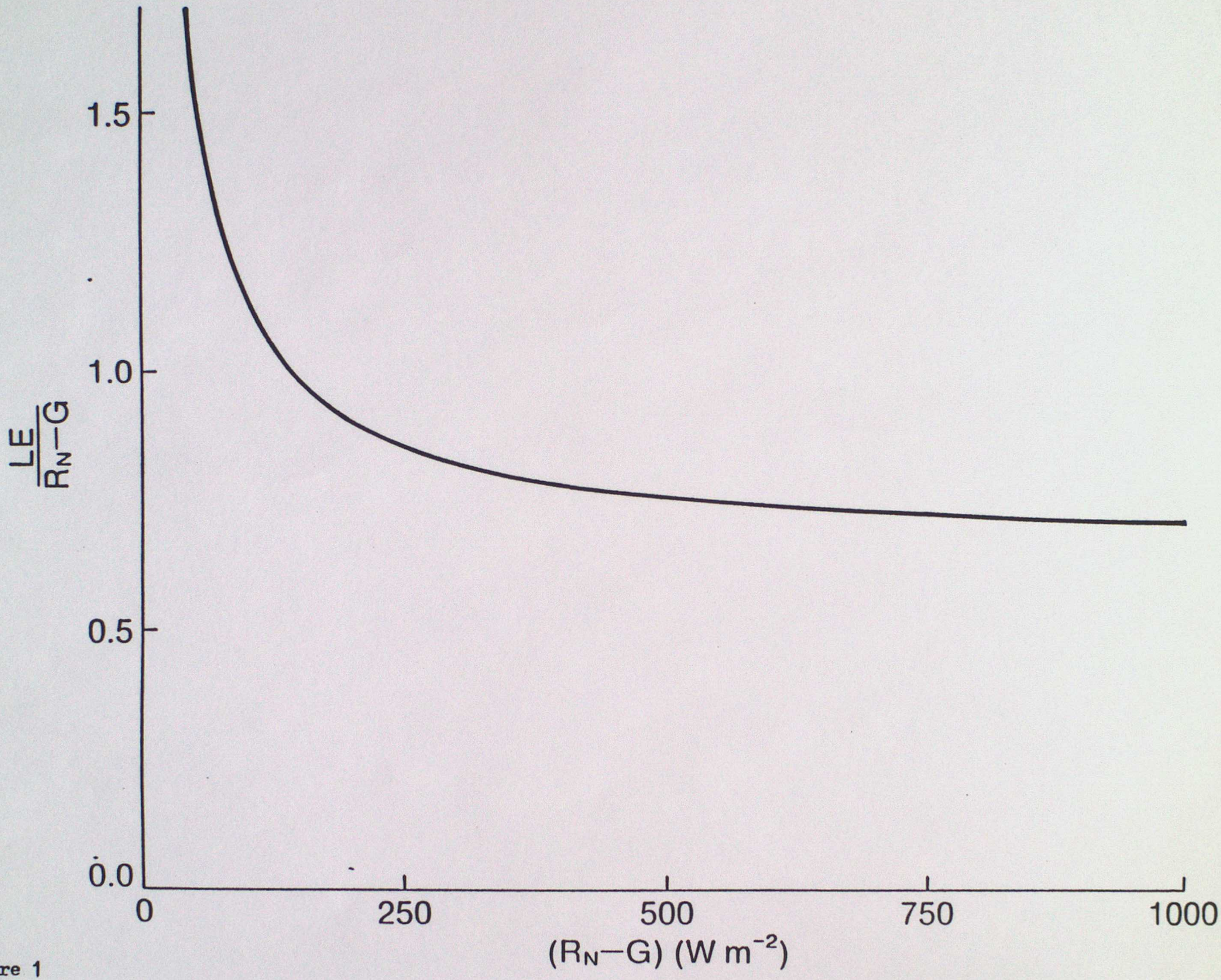


Figure 1

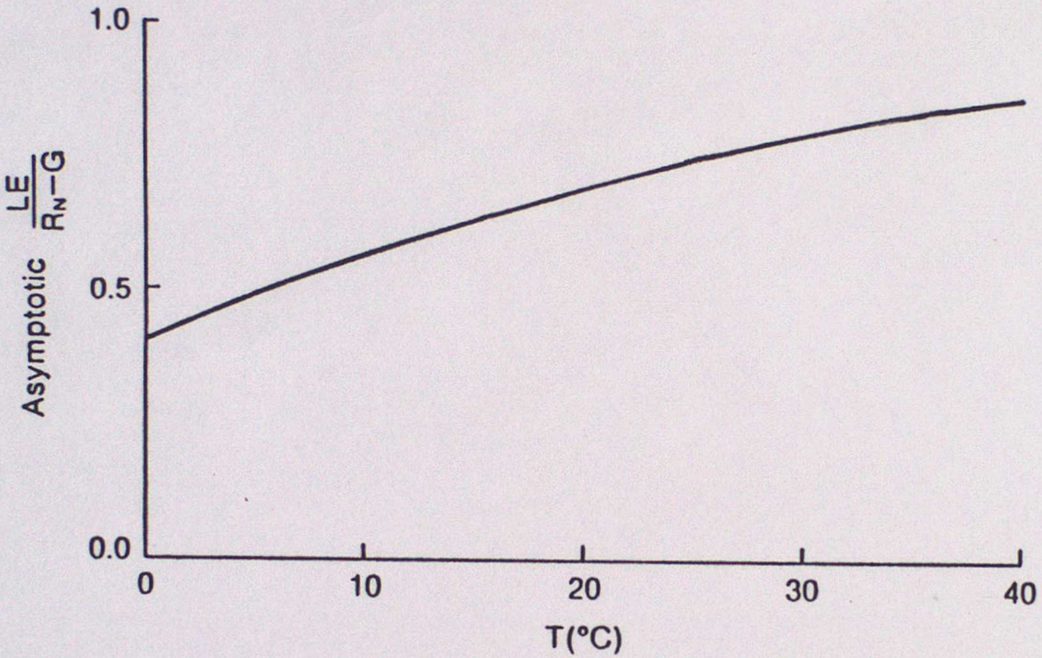


Figure 2

ALBEDO CHANGE: Change in LE as fraction of change in ($R_N - G$)

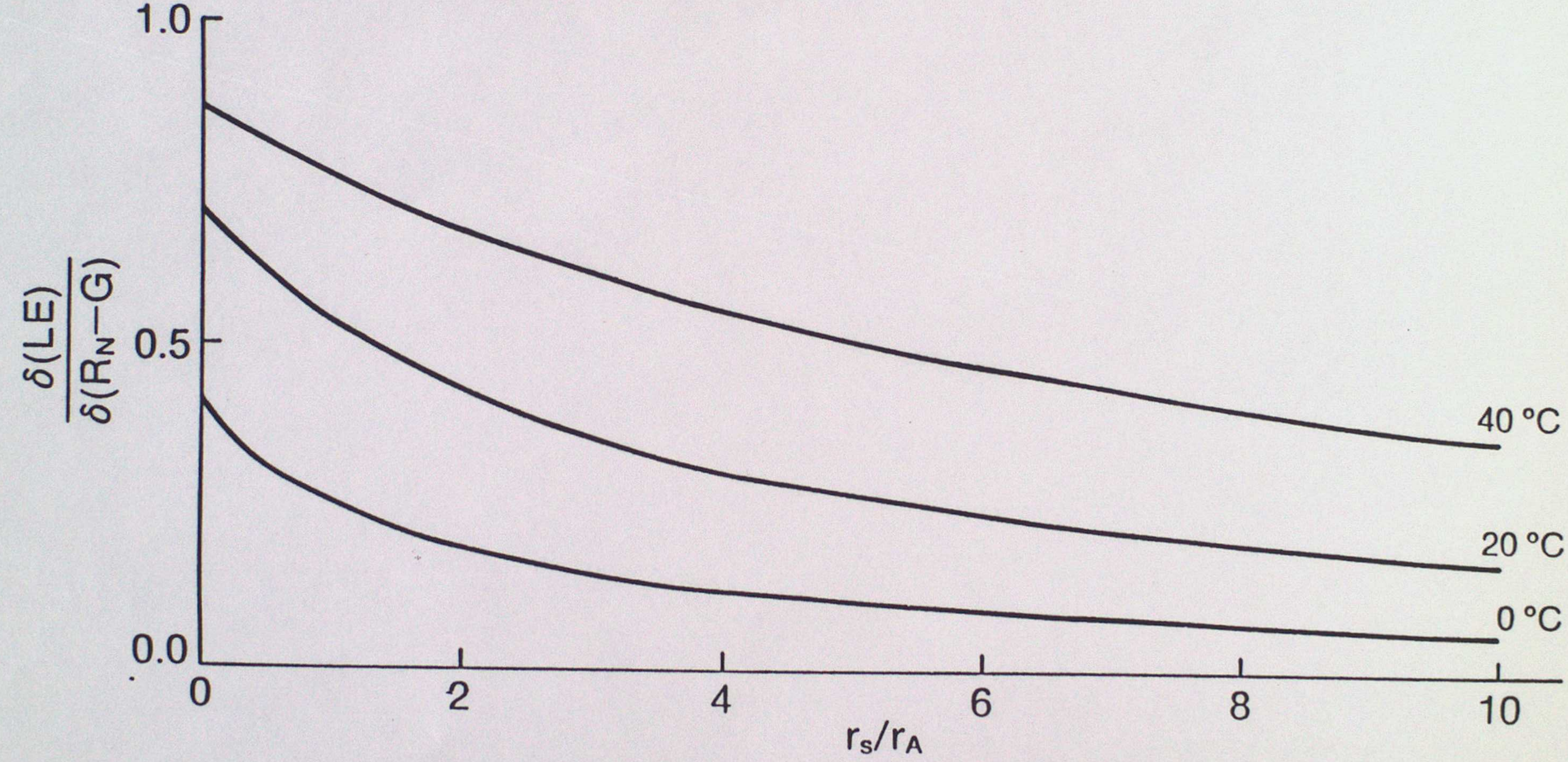
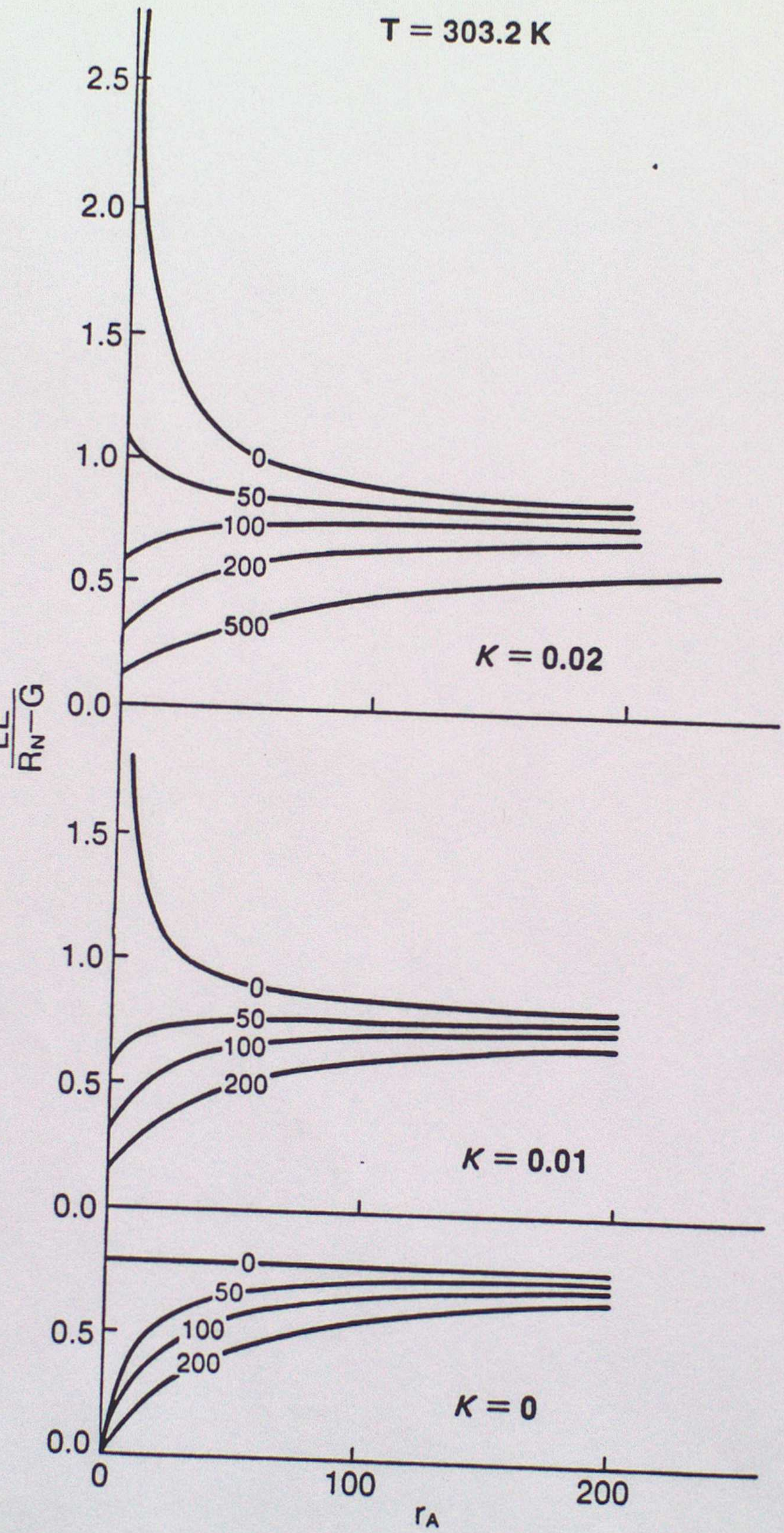


Figure 3



T = 303.2 K
$\frac{LE}{R_N - G}$
2.5
2.0
1.5
1.0
0.5
0.0
0
50
100
200
500
K = 0.02
1.5
1.0
0.5
0.0
0
50
100
200
K = 0.01
0
50
100
200
0.5
0.0
K = 0
0
100
200
r_A

Figure 4 (a)

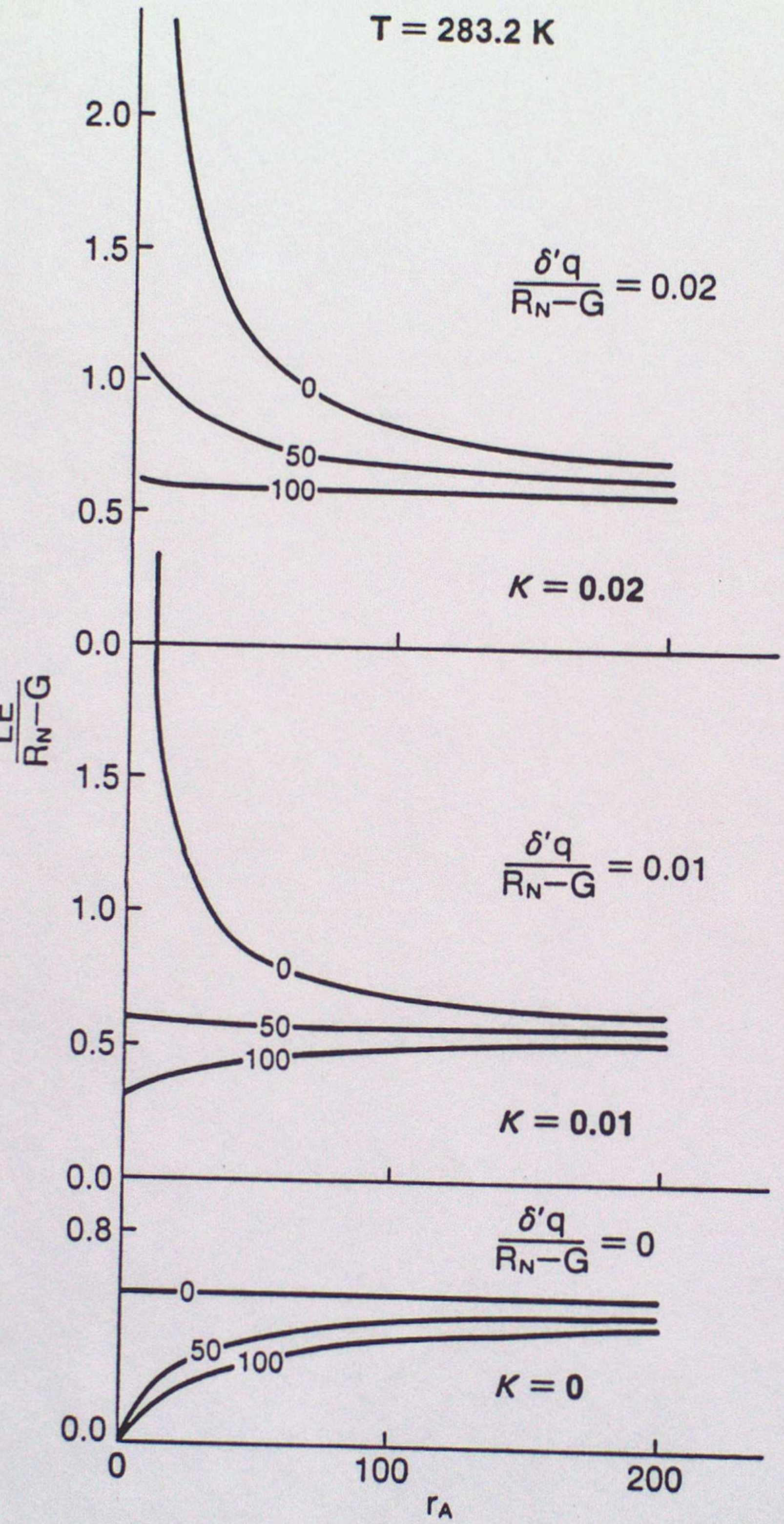

T = 283.2 K
2.0
1.5
1.0
0.5
0.0
δ'q / (R_N−G) = 0.02
0
50
100
K = 0.02
LE / (R_N−G)
1.5
1.0
0.5
0.0
δ'q / (R_N−G) = 0.01
0
50
100
K = 0.01
0.8
0.0
δ'q / (R_N−G) = 0
0
50
100
K = 0
0
100
200
r_A

Figure 4 (b)

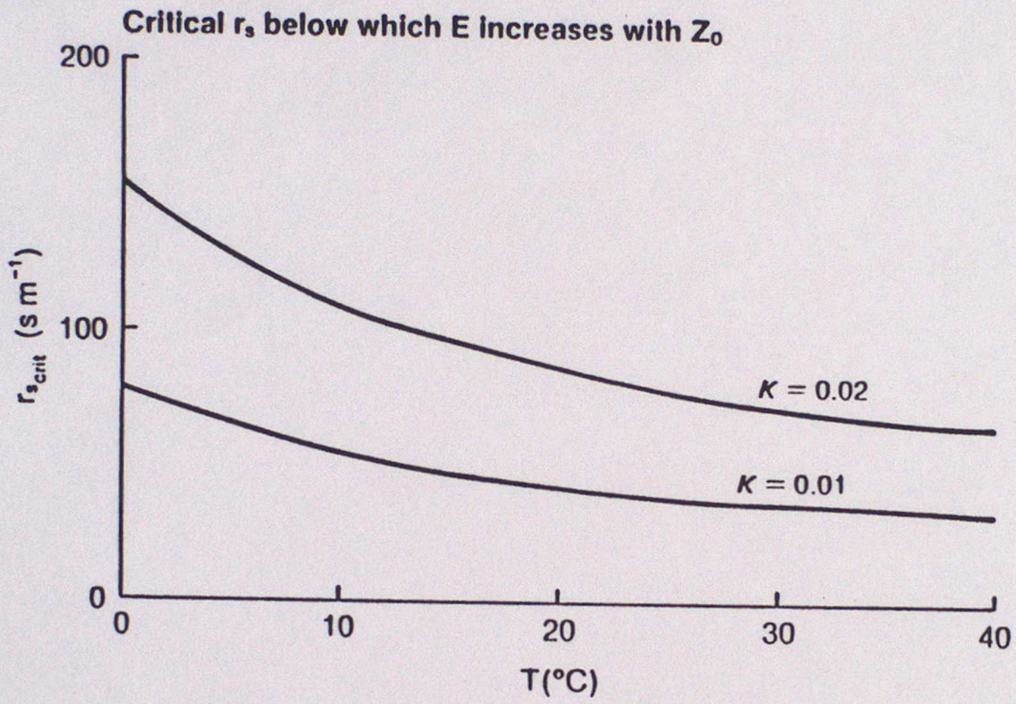

Critical r_s below which E increases with Z_0
200
100
0
$r_{s_{crit}}$ (s m^{-1})
$K = 0.02$
$K = 0.01$
0
10
20
30
40
T(°C)

Figure 5

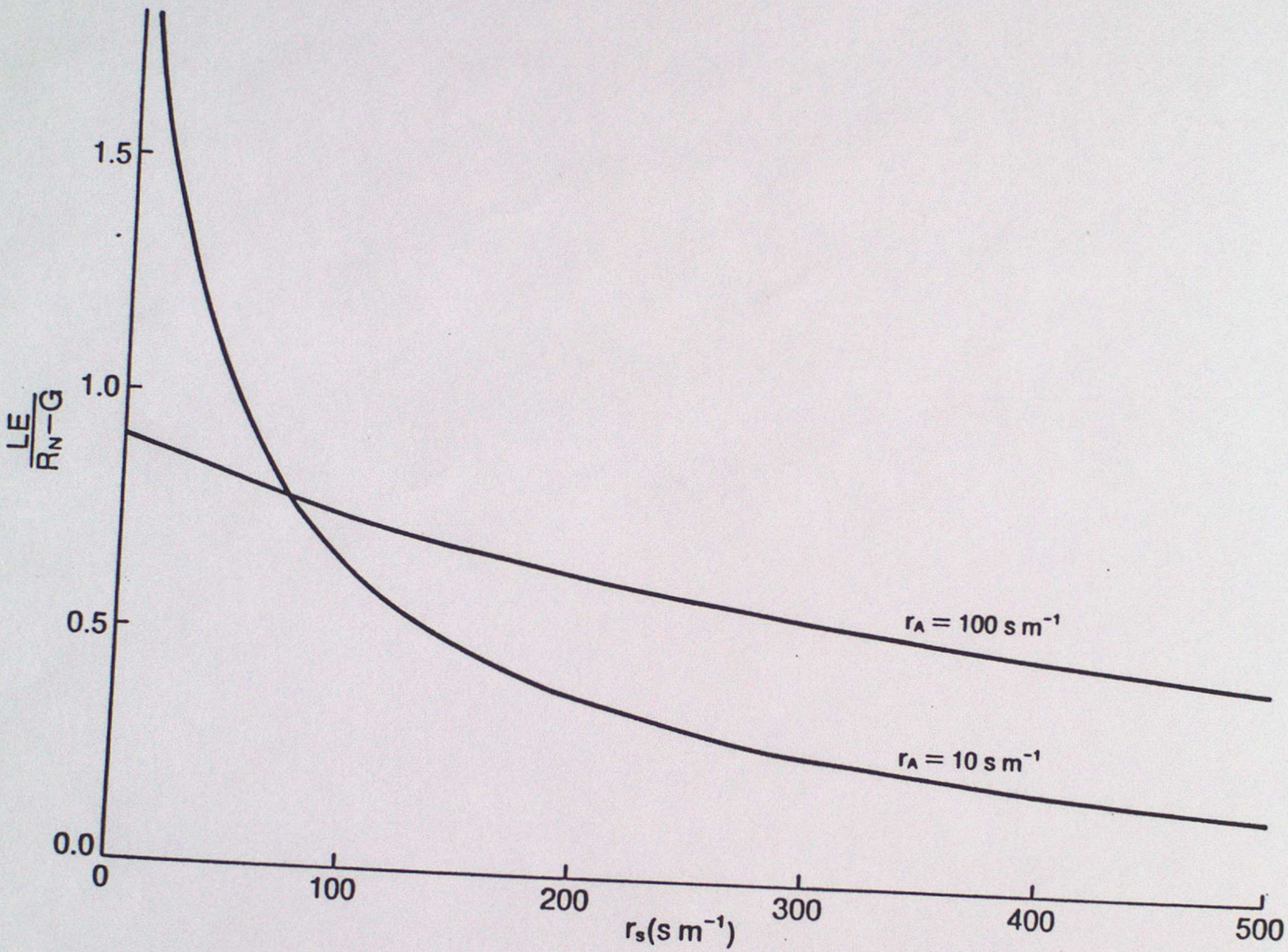


Figure 6